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Features associated with the deglaciation of the upper Saco and Ossipee River basins, northern York and southern Oxford Counties, Maine

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Trip B-1

FEATURES ASSOCIATED WITH THE DEGLACIATION OF THE UPPER SACO AND OSSIPEE RIVER BASINS, NORTHERN YORK AND SOUTHERN OXFORD COUNTIES, MAINE

by

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1.0 INTRODUCTION

The purpose of this field trip is to examine some of the principal features associated with the last deglaciation in an area centered on the Saco and Ossipee River valleys in Oxford and York Counties, Maine. (See Figure 1). The area is located in the New England Upland physiographic province (Fenneman, p. 358) southeast of the White Mountains, and constitutes a physiographic boundary between the seaboard lowland and the mountains. It also marks the limit of Late Wisconsin glaciomarine sediments in the Saco basin (See Figure 2). Because of its locale, this area lends itself to the field application of some of the current working hypotheses concerning the last deglaciation. While none of the features identified in the area is unique to New England, or even to Maine, many of the associations of landforms and deposits are uncommon, and therefore invite discussions of a specific as well as a regional nature.

1.1 ACKNOWLEDGMENTS

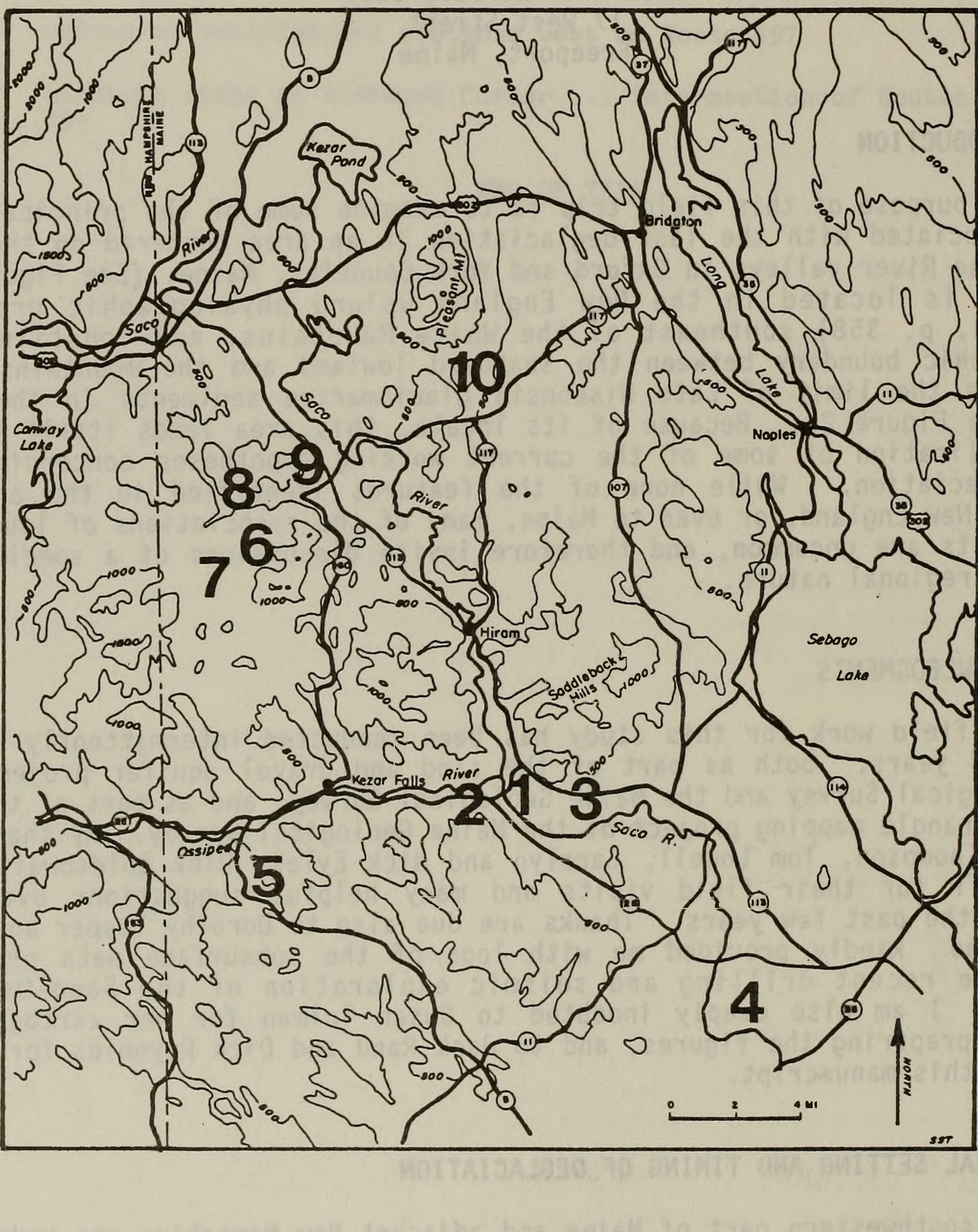
The field work for this study has been conducted intermittently during the last 4 years, both as part of the sand and gravel aquifer project for U.S. Geological Survey and the Maine Geological Survey, and as part of the on-going quadrangle mapping project of the Maine Geological Survey. My thanks go to Woody Thompson, Tom Lowell, Carolyn and Nick Eyles, Dick Goldthwait, and Carl Koteff for their field visits and many helpful suggestions over the course of the past few years. Thanks are due also to Dorothy Tepper and Andy Tolman, who kindly provided me with logs of the subsurface data produced during the recent drilling and seismic exploration of the Saco-Ossipee aquifers. I am also deeply indebted to Susan Tolman for her cartographic skills in preparing the figures, and to Jack Rand and Dick Reynolds for their review of this manuscript.

2.0 GENERAL SETTING AND TIMING OF DEGLACIATION

The southwestern part of Maine and adjacent New Hampshire are underlain chiefly by easily-weathered phaneritic plutonic rocks. In field mapping, this means that striation data are very difficult to obtain at outcrop. Because the stones derived from the local rocks are usually subrounded to rounded,

FIGURE 1

TRIP LOCATION MAP



even in diamicton exposure, fabric data are also difficult to obtain. For these reasons, little is known of the pre-latest Wisconsin history of the region. The dominant striation direction throughout the region is S15-30E, with a much fainter, younger southerly direction at times preserved. Although logic, and inference from other areas that these directions are both Late Wisconsin in age, this conclusion is equivocal. It is because of a lack of information regarding glaciation that this trip concentrates on the last deglaciation, the features of which are at least well preserved in the area. Figure 2 is a generalized map showing the distribution of the major deglaciation deposits in the region.

In the seaboard lowland of southwestern Maine, deglaciation was accompanied by a period of marine inundation. Throughout the area submerged, field evidence indicates that ice was marine-based, and that submergence and ice retreat were coeval. Ice-marginal configurations can be approximated in the area of marine submergence by following the traces of the abundant DeGeer moraines (Smith, 1982). Field evidence also indicates quite clearly that the ice was internally active while the margin stood in the marine environment (Thompson, 1982; Smith, 1982; Lepage, 1979; Thompson and Smith, 1981).

Smith (1985) established a tentative chronology of the last deglaciation of the seaboard lowland of Maine based on dates determined from seaweed, marine shells, and organic sediments collected throughout the area. The principal features of this chronology, as they relate to southwestern Maine are: 1) by 13,800 years ago, the retreating ice margin lay along the present coastline in the vicinity of Great Hill in Kennebunk, based on a date on sediments between what are interpreted to be 2 tills (Smith, 1985, p. 34). The ice margin fluctuated in a relatively restricted belt about this position for a period of roughly 600 years, as indicated by a date of 13,200 on shells in deformed marine sediments at the Kennebunk dump; 2) by about 12,500 years ago, marginal conditions lay inland of what field mapping and paleo-sea level determinations indicate was the "marine limit"; 3) by 11,500 years, the sea had regressed completely from the seaboard lowland.

This chronology is not without dispute. One of the more important questions concerns the nature of the Great Hill date. If the sediment overlying the dated fossiliferous marine silts is not in fact till, but a nonglacial diamicton, as has been suggested (B.D. Stone, 1986, personal communication), then the date would be a <u>minimum</u> date for <u>deglaciation</u>, and not a <u>maximum</u> date for the <u>presence of ice</u>. Since most of the dates acquired in the seaboard lowland are minimum dates, it is conceivable that ice marginal conditions were experienced along the present coastline much earlier than the dates would seem to indicate.

The timing of deglaciation of the area inland of the marine limit in southwestern Maine and adjacent New Hampshire is not currently established due to the lack of dates which can be tied directly to the presence of ice. Davis and Jacobson (1985, p. 358), in a regional paleoecological study, hypothesize that by 13,000 years B.P., western as well as coastal Maine, all of Vermont and all of New Hampshire were ice free, leaving a restricted ice mass in northern and central Maine.

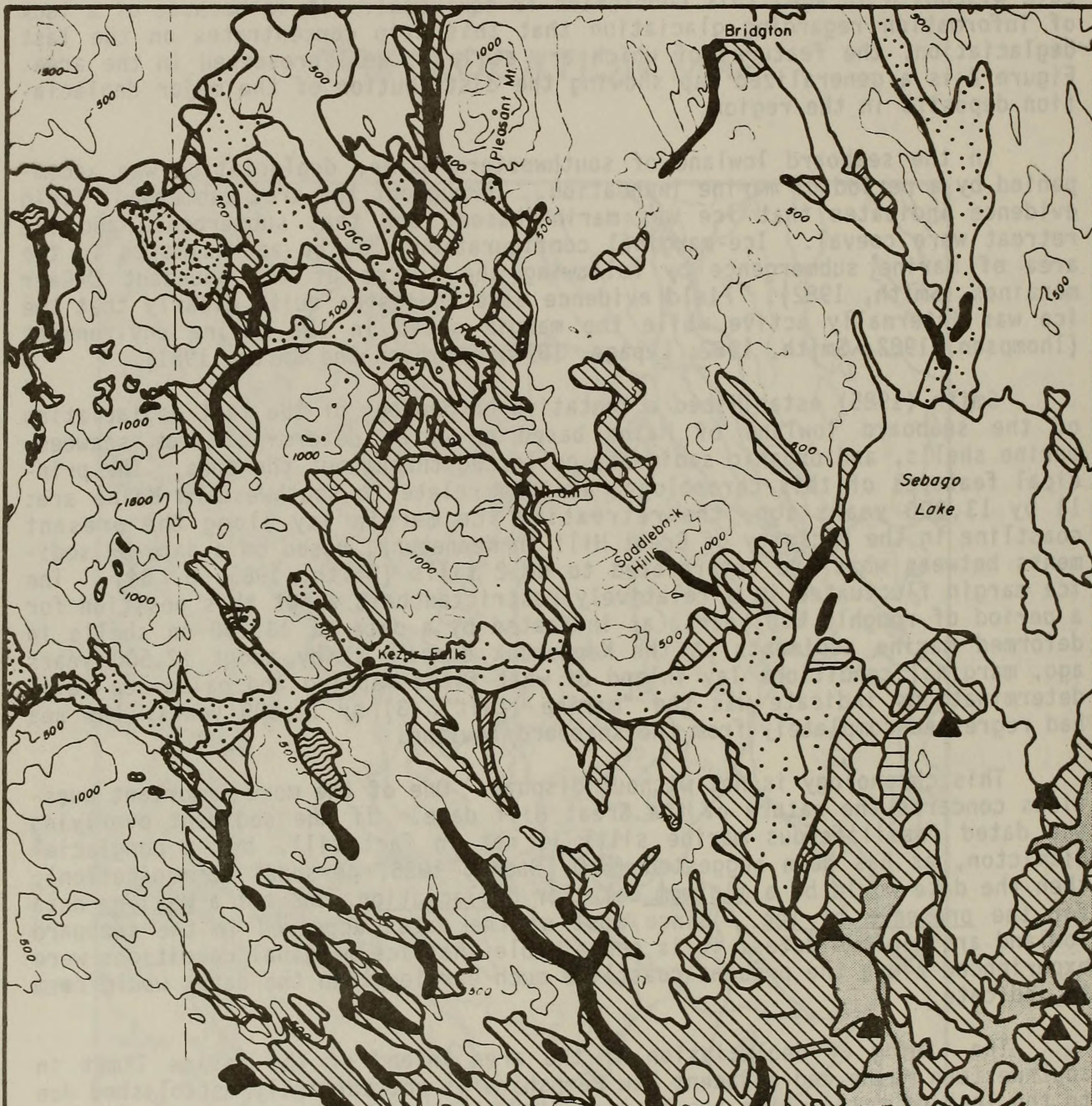

Bridgton
Pleasant Mt
Soco
Hiram
Saddleback Hills
Sebago Lake
Kezar Falls
1000
500
1500
NORTH
0 2 4 MI
SST

FIGURE 2

GENERALIZED DISTRIBUTION OF PRINCIPAL SURFICIAL FACIES

LEGEND

PROGLACIAL FACIES: Includes outwash deltas, collapsed outwash, valley train deposits, and outwash fans.

GLACIOMARINE FACIES: Includes silts, sands, and clays of subaqueous glaciomarine affinities.

SUBAQUEOUS GLACIAL LAKE FACIES: Restricted to lake bottom deposits.

PROXIMAL GLACIAL STREAM FACIES: Includes kames, kame plains, kame plateaux, kame fields, kame terraces, kame deltas, and cross-valley crevasse-fill. Does not include long-valley englacial facies.

ICE CHANNEL FACIES: Includes the principal esker systems, as distinct from ridges of sediment between kettles, and from crevasse-fill.

ICE DISINTEGRATION FACIES: Restricted to deposits composed chiefly of bouldery diamicton, with very minor volumes of stratified sediment, and having a highly irregular morphology.

SUBMARINE ICE MARGINAL FACIES: Includes the larger DeGeer moraines and subaqueous fans.

TERRESTRIAL MORAINE FACIES: Applies to so-called "ribbed moraine" of uncertain genesis.

UNDIFFERENTIATED DIAMICTON FACIES: Includes deposits of till, and all other diamicton facies not included elsewhere.

SPECIAL FEATURES

DEGEER MORAINES: Locations and numbers stylized.

EXPOSURES OF TOPSET-FORESET CONTACTS IN GLACIOMARINE DELTAS

Sources: Smith (1977a, 1977b), Thompson (in press, 1976a, 1976b), Thompson and Borns (1985), Newton (1974), and Holland, (unpub. mapping).

2.1 THE STYLE OF DEGLACIATION IN SOUTHWESTERN MAINE

"...I have long since learned that glacial rivers bear careful watching. Their deceitfulness is well exhibited..."(Stone, 1899, p.252). The style of deglaciation of southwestern Maine and adjacent New Hampshire landward of the marine limit raises questions about which there little consensus exists. Essentially similar field data have been invoked at different times and by different people as evidence for either areal stagnation, or for progressive, active retreat. One wonders whether at least part of the controversy could be avoided if everyone first agreed upon a meaningful definition of the terms "active" and "stagnant" ice for field use in interpreting ancient deglaciation environments. Such a definition would include a set of recognizable field criteria which could be linked empirically to certain ranges of ice flow velocity through observation of existing glaciers in dynamically equivalent settings. To my knowledge, the only attempt at such a definition as applied to northern New England is that proposed by Goldthwait and Mickelson (1982), using an analogy between the White Mountains of New Hampshire and the Glacier Bay area of Alaska. Using their example, I have assumed a flow velocity of 10 meters/yr to distinguish between "stagnant" and "active" ice. I have also assumed that "dead" or "disintegrated" ice implies velocities approaching or equal to 0 - like an ice cube resting on a table-top.

By way of introduction, it is perhaps more appropriate to present the question of deglaciation style of inland southwestern Maine and adjacent New Hampshire in terms of whether ice retreat involved principally vertical "retreat" (top to bottom), or horizontal retreat (front to back). Stated this way the problem becomes more workable, and there are several lines of background information which are applicable and worthy of consideration: 1) radiometric dates; 2) field geology in the highland areas; 3) field geology from the seaboard lowland; 4) field geology along the larger river valleys above the marine limit; and 5) theoretical modeling:

1) There are numerous dates from high elevations in the mountains of New Hampshire and Maine (Spear, 1981; Davis and Jacobson, 1985; Borns and Calkin, 1977). The dates closest to the trip area are from Deer Lake Bog and Lake of the Clouds (Spear, 1981), in the White Mountains of New Hampshire. The dates for these localities indicate that the high peaks of the Presidential Range were ice free by between 13,000 and 14,000 years B.P.

If the date of 13,800 from Great Hill in Kennebunk is accepted as indicating ice along the coast at that time, and if the dates from the high ponds in the White Mountains are accepted, the concept of a "thinning pancake" (Lowell, 1983) is compelled. Even without the Great Hill date, the Kennebunk dump date still puts ice along the present coastline at a time when at least part of the White Mountains has been deglaciated.

2) Recent work by Gerath, Fowler, and Haselton (1985), Gerath and Fowler (1982), Davis (1986), and Goldthwait and Mickelson (1982) has reaffirmed the interpretation by Goldthwait (1970) that ice thinned over the White Mountains fairly early during deglaciation, and that no reactivation of the White Mountain cirques occurred during "post-Laurentide" time. A similar conclusion was reached by Borns and Calkin (1977) for the Boundary Mountains of Maine, and by Davis (1983) for Mt. Katahdin.

3) There is little question from field evidence that the retreat of ice through the seaboard lowland of Maine was progressive, and involved ice which retained sufficient internal activity to produce an abundant sediment supply, to construct end moraines, and to deform sediments (Smith, 1982, 1985; Thompson, 1982).

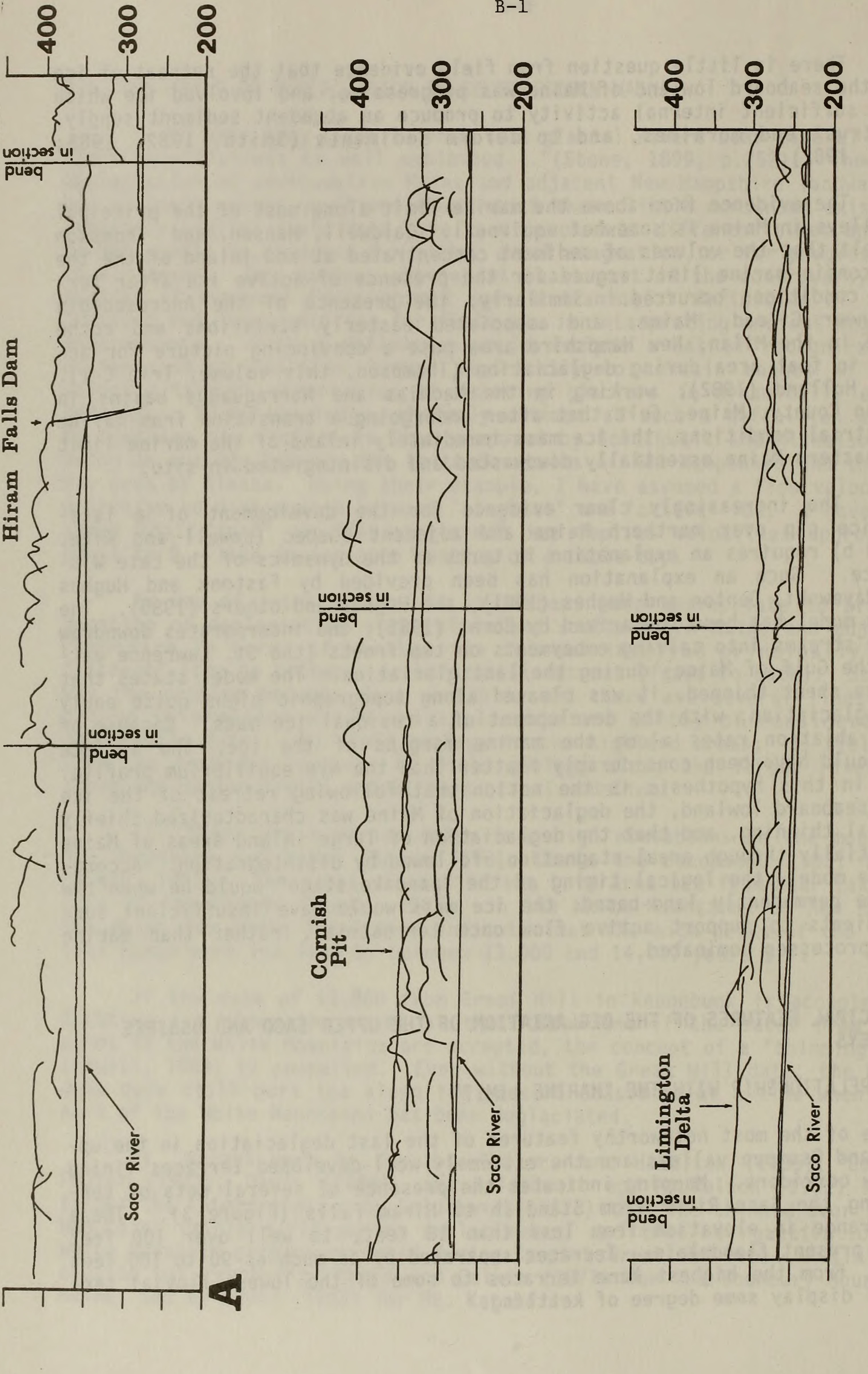
4) The evidence from above the marine limit along most of the principal river valleys in Maine is somewhat equivocal. Caldwell, Hanson, and Thompson (1985) felt that the volumes of sediment concentrated at and inland of the the Late Wisconsin marine limit argued for the presence of active ice after terrestrial conditions occurred. Similarly, the presence of the Androscoggin moraine near Gilead, Maine, and associated easterly striations and roche moutonnee in the Milan, New Hampshire area pose a convincing picture for active ice in that area during deglaciation (Thompson, this volume, Trip C-1). However, Holland (1982), working in the Machias and Narraguagus basins in Washington County, Maine, felt that after undergoing a transition from marine to terrestrial conditions, the ice mass immediately inland of the marine limit in southeastern Maine essentially downwasted and disintegrated in situ.

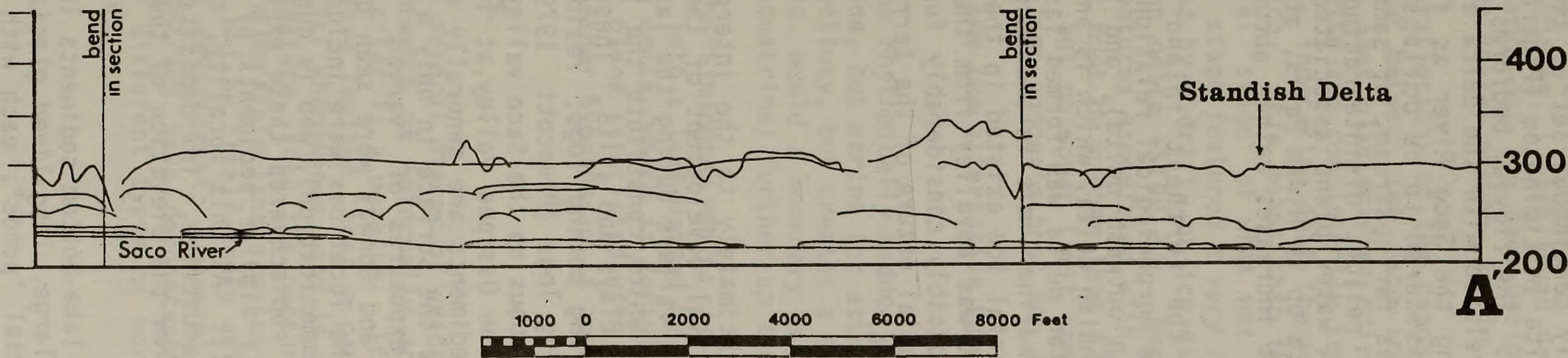
5) The increasingly clear evidence for the development of a late-glacial ice cap over northern Maine and adjacent Quebec (Lowell and Kite, 1986a and b) requires an explanation in terms of the dynamics of the Late Wisconsin ice. Such an explanation has been provided by Fastook and Hughes (1982), Mayewski, Denton and Hughes (1981), and Hughes and others (1985). The resulting model has been summarized by Borns (1985), and incorporates downdraw along ice streams into calving embayments on two fronts (the St. Lawrence valley and the Gulf of Maine) during the last glaciation. The model states that as the ice sheet thinned, it was cleaved along topographic highs quite early during deglaciation, with the development of a residual ice mass. Because of the high ablation rates along the marine margins of the ice, the surface profile would have been considerably flatter than the Nye equilibrium profile. Implicit in this hypothesis is the notion that following retreat of the ice from the seaboard lowland, the deglaciation of Maine was characterized chiefly by vertical thinning, and that the deglaciation of large inland areas of Maine was essentially through areal stagnation, followed by disintegration. According to the model, the logical timing of the "pancake stage" would be when the ice became permanently land-based; the ice mass would have insufficient surface gradients to support active flow once terrestrial, rather than marine ablation processes dominated.

3.0 PRINCIPAL FEATURES OF THE DEGLACIATION OF THE UPPER SACO AND OSSIPEE VALLEYS

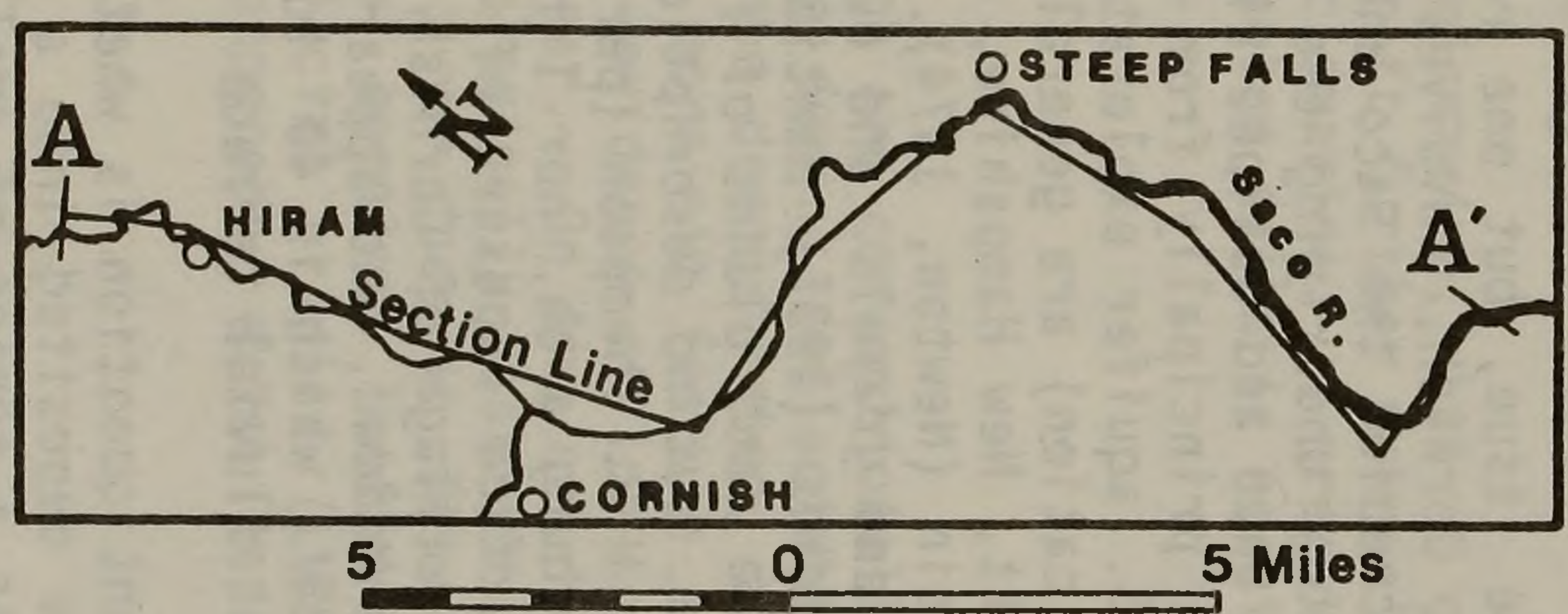
3.1 THE RELATIONSHIP WITH THE "MARINE LIMIT"

Some of the most noteworthy features of the last deglaciation in the upper Saco and Ossipee valleys are the extremely well-developed terraces lining the valley corridors. Mapping indicates the presence of several sets of terraces along the Saco River from Standish to Hiram Falls (Figure 3). These terraces range in elevation from less than 10 feet, to well over 100 feet above the present floodplain. Terraces separated by as much as 90 to 100 feet of relief, from the highest kame terraces to some of the lowest fluvial terraces, all display some degree of kettling.


A
Hiram Falls Dam
Cornish Pit
Limington Delta
Saco River
Saco River
Saco River
bend in section
bend in section
bend in section
bend in section
bend in section
400
300
200
400
300
200
400
300
200



SECTION LOCATION MAP



NOTES:

1) Vertical scale on sections is in feet above NGVD.

2) Profiles constructed by orthogonal projection of the highest surfaces of stratified deposits on either side of the section line.

3) Sections read from north to south, from the top to bottom in the figure.

FIGURE 3

TOPOGRAPHIC PROFILES OF STRATIFIED DEPOSITS IN THE SACO RIVER VALLEY

STOP 1 is an exposure in the most prominent terrace set in the Cornish area. The feature is part of a relatively continuous set of terraces which grade into an apparently glaciomarine delta in Limington (Figures 2 and 3), over which we will pass en route to **STOP 3**. The fluvial bedforms displayed in this pit are typical of the terraces correlated with the delta along the Ossipee River into New Hampshire, and along the Saco River as far as Hiram Falls. The deposits as a whole are considered to be collapsed outwash, resulting from the deposition, from sources far upstream, of sand and gravel over a thin, discontinuous assemblage of relict, disintegrating ice masses lying in the Saco and Ossipee valleys. A water plane associated with these terraces may be easily drawn from the delta up to Hiram Falls, at which point the surface elevations of stratified drift increases rather abruptly (Figure 3).

The delta at Limington is the furthest inland of all the glaciomarine deltas in the Saco basin. The morphological relationship between the distal part of the Limington delta (an outwash delta), and the proximal part of the next delta downstream (the Standish delta, which is an ice-contact delta; **STOP 4**) indicates that the Limington delta was formed last, and suggests that it therefore marks the maximum marine limit in the Saco basin. If a zone of relatively thick, continuous glacial ice existed in the area at the time of the maximum marine submergence, it was upvalley from the delta some distance- at least to Hiram Falls, and possibly considerably further north. This is indicated by the fluvial and proglacial nature of the terrace deposits and the association with the Limington delta.

3.2 THE DEVELOPMENT OF GLACIAL LAKES

A separate issue, but one related intimately to the interpretation of the terraces at Cornish, involves the initial development of the proglacial lake(s) which occupied the Saco and Ossipee valleys during the last deglaciation. Conformably underlying the deposits which constitute the fluvial terraces along the Saco and Ossipee rivers are clays and silts. These sediments, which are known principally from engineering boring logs, residential well logs and U.S.G.S. aquifer exploration boring logs (Prescott, 1979; D.Tepper, personal communication) are generally continuous up the Saco valley from Hiram Falls to Bartlett, New Hampshire, and up the Ossipee valley,at least to the Ossipee Lake basin (Newton, 1974). The sediments are generally quite deeply buried by sands and gravels, and outcrop at the surface in only a few places. Leavitt and Perkins (1935) mention that exposures of "varved" fresh water clays beneath the sand plains of Fryeburg and Conway were seen during their work in the area, but no descriptions of the locations were included in their report. Neither W.B. Thompson (personal communication), who is currently mapping in the Fryeburg area, nor I have yet rediscovered these exposures. By far the best surface exposure of seemingly similar sediments of which I am aware is in the village of Cornish (**STOP 2**). At this locality, the deposits are rhythmically bedded, and appear to be lacustrine. The deposits outcrop at approximately 325', which is 40' to 50' below the elevation of the surface of the superadjacent fluvial terrace.

An important question is whether the lake-bottom sediments in the upper Saco valley were deposited into a single large, long-lived glacial lake, or whether there was a series of smaller glacial "ponds", each with a separate base-level control and a unique water plane. This question is difficult to answer directly; the Saco is a south-draining river, meaning that the only possible dams in the Saco valley would be either relict ice and/or drift, and as of this writing, I have found no irrefutable dams or spillways. However, in certain areas of the Saco-Ossipee corridor, such as near the village of Cornish, a potential answer to the lake problem may be deduced.

The stratigraphy of the deposits at Cornish indicates that in the area of confluence between the Saco and Ossipee Rivers, the lacustrine episode preceded the last glaciomarine episode. The fluvial terraces grading to the Limington delta are consistently kettled (indeed, even lower terraces are often kettled), indicating that the valley was far from ice-free prior to terrace formation. This means that the clays exposed at Cornish were deposited in the initial "stage" of what was ultimately a somewhat larger glacial lake. This initial stage featured a very narrow east-west strip of open water, which was impounded by a drift- or ice dam across the valley 0.2 miles upriver from the Limington delta (**STOP 3**). Because of the lack of deltas, spillways or shoreline features associated with this stage, its water-surface elevation is unknown. It must be higher than 325' (the elevation of the bottom sediments at Cornish), but no upper limit can be found. Presumably, the lack of water-plane indicators suggests that the initial stage of the lake was short-lived.

There are two alternative hypotheses that can also accommodate the field data:

1) The rhythmically-bedded deposits beneath the terrace in Cornish are not lacustrine, but marine (or estuarine). This obviates the need to hunt for a dam and spillway, but creates a separate problem: at what point, and on what genetic basis, would a boundary be established with the clay/silts (which are certainly lacustrine) further up the Saco valley ?

2) The Limington delta is not glaciomarine, but a glaciolacustrine delta, built into a lake dammed by ice or drift further downvalley; perhaps by the Standish delta, which is definitely glaciomarine. A precise determination of the water plane associated with the Limington delta is required in order to determine whether hypothesis is appropriate.

3.3 MORAINES

Of the several pieces of evidence which are conventionally used to document the presence of internally active ice during deglaciation, one of the more sought after is a moraine. If it can be demonstrated, as has been done in the seaboard lowland, that a particular morainic feature or assemblage is an ice-marginal phenomenon, and if there is evidence of ice-shove deformation within the feature, then the picture is clear. On the other hand, the mere presence of a morainic landform (as a morphologically distinct and independent entity) does not by itself imply active ice, nor does it necessarily imply an ice marginal environment. Several types of morainic landforms in the Scandinavian countries have demonstrably subglacial origins (Lundqvist, 1969). Two uniquely different types of morainic topography exist in the upper Saco-Ossipee basin.

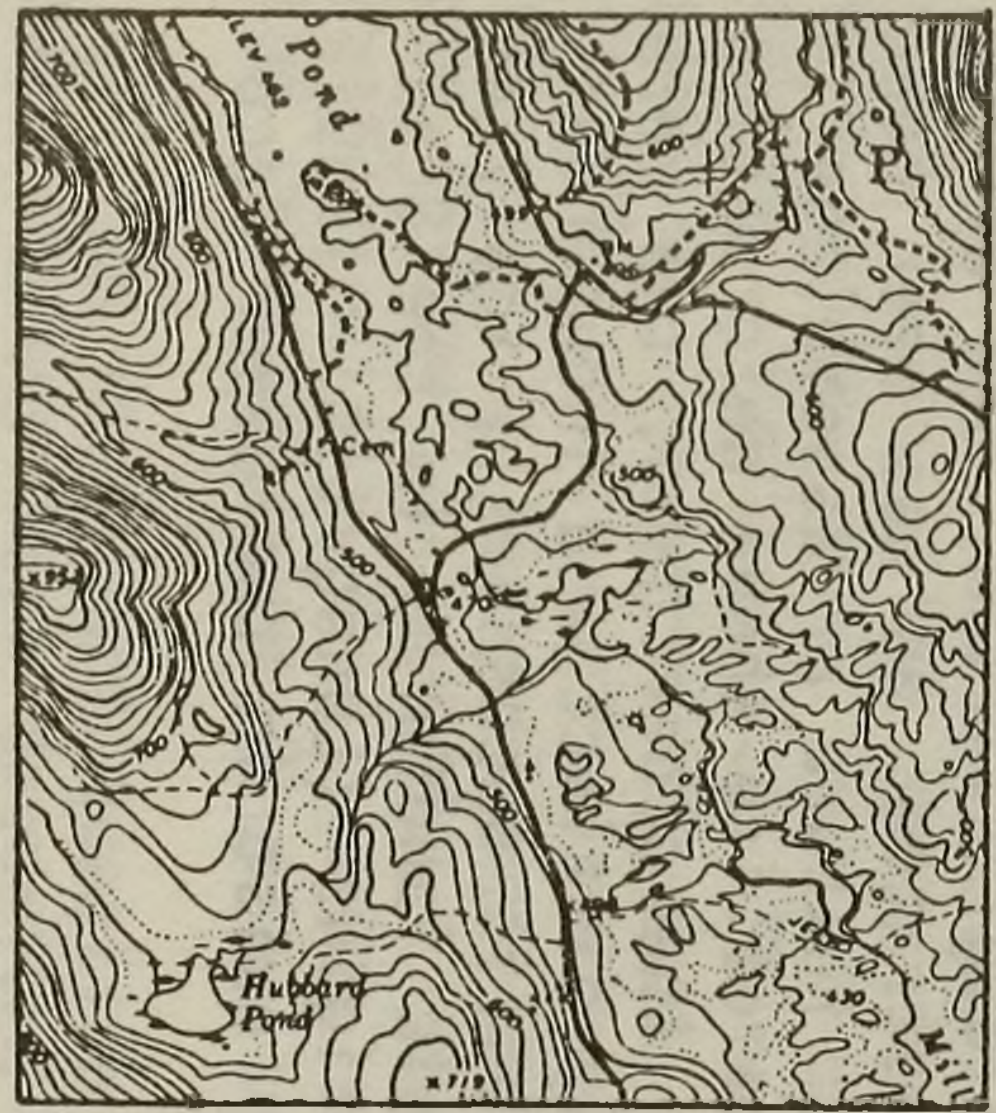


FIGURE 4a

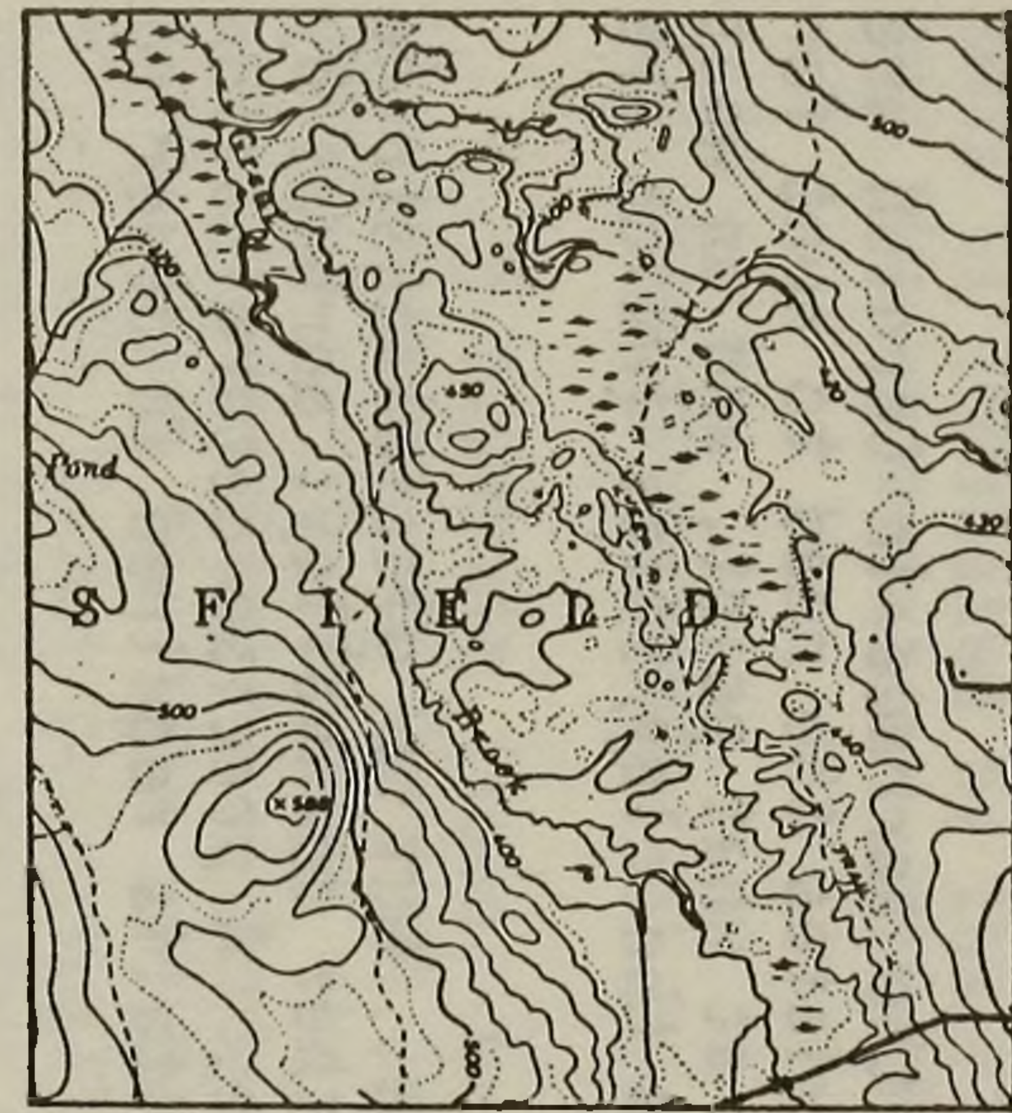


FIGURE 4b

From U.S. Geological Survey Kezar Falls 7.5' Topographic Quadrangle Map

CONTOUR INTERVAL 20 FEET

1000 0 1000 2000 3000 4000 5000 6000 7000 FEET

1 .5 0 1 KILOMETER

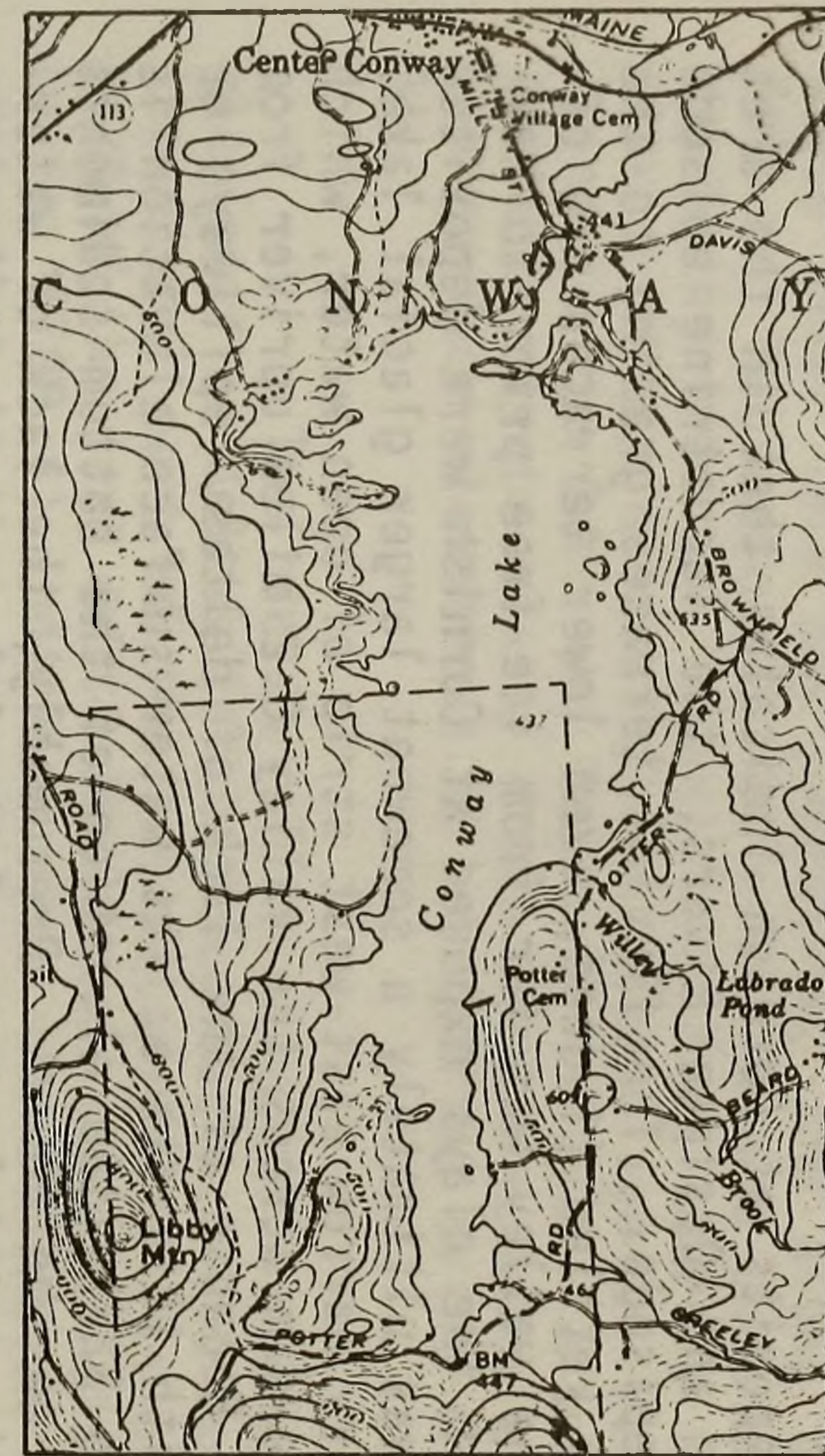
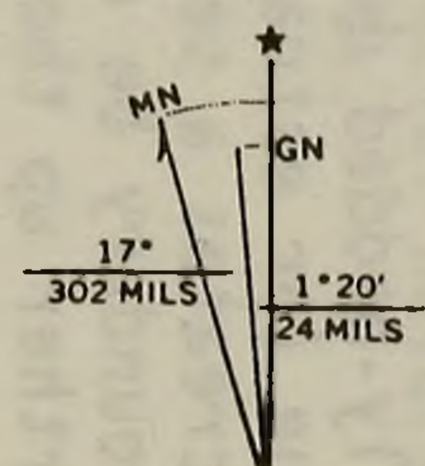


FIGURE 4c



From U.S. Geological Survey Ossipee Lake 15' Topographic Quadrangle Map

3000 0 3000 6000 9000 12000 15000 18000 21000 FEET

1 .5 0 1 2 3 4 5 KILOMETERS

CONTOUR INTERVAL 20 FEET
DATUM IS MEAN SEA LEVEL

FIGURE 4

MORPHOLOGY OF "RIBBED MORAINE"

3.3.1 "RIBBED" MORAINE

The first type includes independent linear ridges, composed exclusively of coarse-grained glacial diamicton. Within the study area, the ridges are found exclusively in the bottoms of second- or third- order tributary valleys whose trends range from N0E-S0W to N30W-S30E, subparallel to what are most likely the 2 most recent directions of glacier flow, as inferred from striations. The length:width ratios of these moraines generally exceed 4:1, but are less than 10:1, making the moraines noticeably "stubbier" than the coastal DeGeer moraines. The long axes are oriented generally normal to the long axes of the containing valleys; the moraines occur exclusively in clusters, or "swarms"; and they have been found in both north-draining and south-draining valleys.

Whether the repetitive morphology of the moraines is an indication that internally active ice was involved in their formation is open to question. Even if active ice was responsible, however, there is still the question of the environment of deposition of the moraines. Preliminary field work indicates that in those cases where proglacial stratified sediment coexists with the moraines in a given valley, it overlaps them on both their distal and proximal faces, suggesting that the moraines might not be marginal features, but submarginal or subglacial. Although it is possible that the overlying proglacial sediments may have been produced sequentially upon retreat of ice to successive marginal positions (each of which would be marked by a moraine "line"), the uniform texture of the stratified deposits from the proximal and distal sides of a single moraine would seem to argue against this interpretation. If the moraines are subglacial, then whether or not they were formed by active ice becomes a moot point when considering them as indicators of deglaciation style; they could have formed very early during deglaciation, when the ice was still relatively thick.

Figure 4 shows the appearance of several of these moraine localities on both the 15-minute and 7.5-minute quadrangle map scales. Superficially similar moraines have been identified elsewhere in Maine (Lowell, 1981; Caldwell, 1976; Holland, 1983; Thompson and Borns, 1985), but a consistent and uniform interpretation of their genesis has not yet appeared. As Caldwell, Hanson, and Thompson (1985) observe, the regional distribution of the features appears to be restricted, as it is in this area, to terranes of phaneritic plutonic rock. The features also bear at least an initial resemblance to the so-called "Storso moraine" described by Lundqvist (1981) in Scandinavia. Because of the current lack of understanding concerning the origins of these features, the reminology presently used to map them in Maine is simply "ribbed moraine".

STOP 5 is in one of the most accessible and best exposed of the ribbed moraine localities in the region. It is hoped that the evidence presented at this stop will generate discussion, because the interpretation of these moraines is one of the keys to forming a working hypothesis concerning the mode of deglaciation of the region.

3.3.2 "ICE-DISINTEGRATION MORAINE"

The second type of morainic deposits exhibit an extremely hummocky, essentially random topography, and is generally found on valley sides above

FIGURE 5a

FIGURE 5b

Topography From U.S. Geological Survey Kezar Falls 7.5' Quadrangle Map

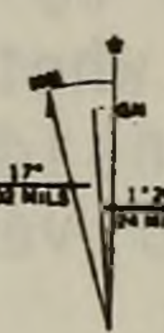
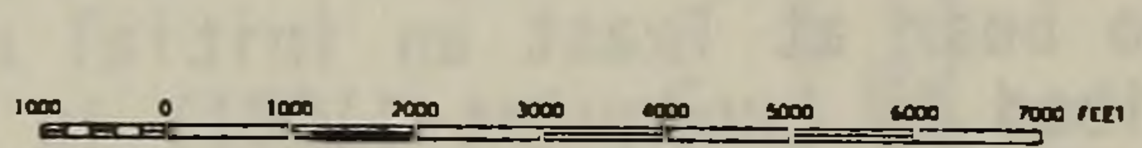


FIGURE 5c

Topography From U.S. Geological Survey Hiram 7.5' Quadrangle Map

FIGURE 5d

Topography From U.S. Geological Survey Cornish 7.5' Quadrangle Map



LEGEND

BOULDERY DIAMICTON FACIES

PROXIMAL GLACIAL STREAM FACIES

DISTAL GLACIAL STREAM FACIES

MELTWATER CHANNELS

FIGURE 5

ICE DISINTEGRATION FEATURES

the locally highest stratified deposits. The deposits are composed largely of sandy, bouldery diamicton, but also contain minor intervening "pockets" of stratified sediment, exposed both at the surface and in cuts. There is almost always a close spatial relationship between the deposits and lateral meltwater channels (see Figure 5). In many instances, there is a deposit of continuous stratified sediment which extends downvalley away from the morainic debris. Most of the stratified deposits exhibit some degree of collapse. The occurrences of stratified sediment in association with the morainic debris appear to be restricted to south-draining tributary valleys.

I have classified the deposits tentatively as "ice-disintegration moraine" because of the texture of the sediment; the position of the deposits relative to the principal deposits of stratified drift in a given valley (as contrasted to the relatively minor volumes of stratified drift included within the morainic mass); and the close association with well-developed meltwater channels, and the similarity to features described by others elsewhere (Boulton, 1967; Gray and Lowe, 1977; and Sissons and Sutherland, 1976) as representing superglacial deposition in an essentially dead-ice environment.

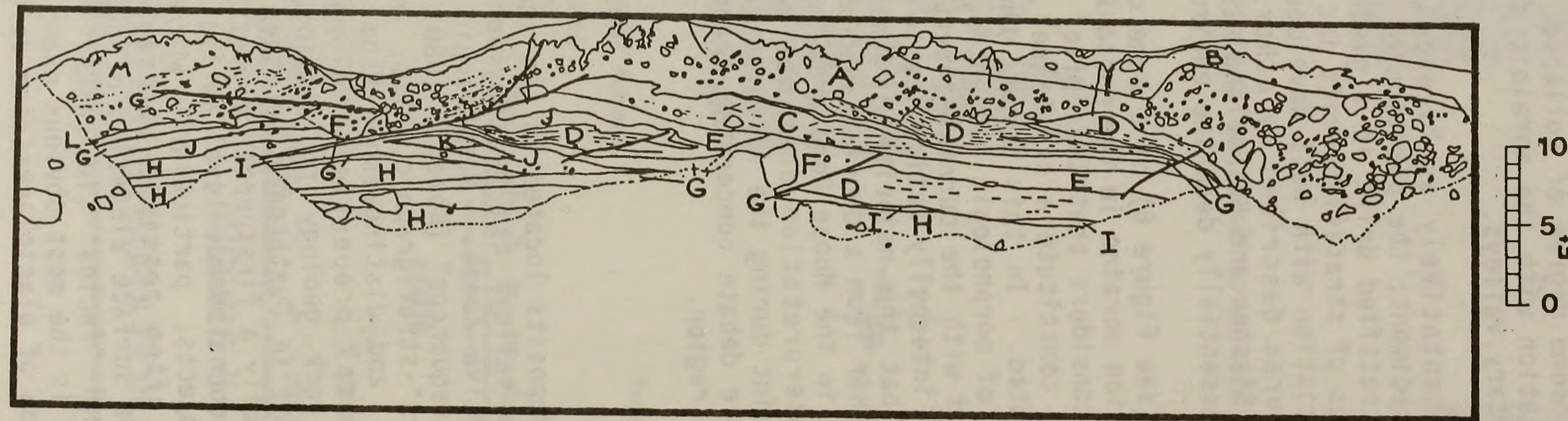
STOP 10 will examine an exposure (see Figure 6) of a kame complex spatially associated with an ice-disintegration moraine in the Moose Pond area of Denmark. My current working hypothesis considers that the "pile" of morainic debris, and the associated kame complex, constitute a single set of deposits that are genetically and temporally related. In this sense, they may constitute a special variation on the theme of morphologic sequences, as defined by Koteff (1974). However, as contrasted with the current interpretation of sequences, which requires the presence of internally active ice to produce the sediment, this interpretation implies that the majority of the stratified sediment has been derived indirectly: either from a debris-laden disintegrating ice mass (which in this case resided in the Moose Pond area), or from the morainic materials themselves. The interpretation does not require the presence of active glacial ice at any point during the formation of the complex. I expect and welcome the inevitable debate concerning these and other similar hummocky morainic deposits in the region.

3.4 MORPHOLOGIC SEQUENCES

Several of the sets of stratified deposits located in the smaller highland basins between the Saco and Ossipee valleys can be considered as morphologic sequences in a strictly descriptive sense. Morphologic sequences, or, in the more euphonious form, "morphosequences" (Koteff, 1980) constitute the surficial mapping equivalents of "rock"-stratigraphic units, in that they can be identified in the field, mapped, and distinguished from other such units on the basis of observable physical properties. Along with the properties conventionally used in hard-rock geology to distinguish rock-stratigraphic units, the geologist working in Quaternary terranes also uses morphology. In fact, morphology is generally a first-order property which is often the only field criteria used in reconnaissance-level surficial mapping projects. In more detailed mapping projects, particularly where abundant topographic data are available, it is often possible to establish morphostratigraphic units on the basis of the surface elevation of deposits, as well as on the basis of the morphological "facies" (ie, esker vs. valley train). The correlation of river terraces is the most obvious example of such an exercise. Similarly, the differentiation of glaciolacustrine deltas on the

FIGURE 6

PICTORIAL SECTION OF SEDIMENTS EXPOSED IN THE MOOSE POND KAME COMPLEX



LEGEND

A Sand boulder gravel, poorly to moderately sorted, crudely stratified; boulders subangular to subrounded.

B Gravelly fine to medium sand, bimodal texture, gravel to small cobble size, massive to crudely stratified.

C Gravelly coarse sand, very well sorted with openwork texture, gravel chiefly pebble-sized, planar bedding.

D Medium sand, very well sorted, planar bedding.

E Silty sandy pebble gravel to silty gravelly sand, moderately well sorted, tabular bedding.

F Gravelly silty sand, poorly sorted, massive diamicton.

G Gravelly silty sand, poorly sorted, matrix-supported, crudely stratified diamicton.

H Gravelly fine sand. Gravel chiefly pebble-sized, crudely laminated. Contains thin (<0.1') interbeds of stratified diamicton similar to unit G.

I Slightly gravelly, silty fine to coarse sand. Moderately well sorted, stratified, gravel chiefly pebble-sized.

J Silty sandy gravel, moderately poorly sorted, gravel clasts from small pebble to large cobble size, clast-supported.

K Pebble gravel, very well sorted with openwork texture, cross-stratification.

L Sandy gravel, moderately well sorted, gravel clasts range from pebble to cobble size.

M Gravelly sand, complexly interbedded with medium sand, silty sand, coarse sand, pebble gravel, and stratified sandy diamicton. Contains clasts to small boulder size.

NOTES

1) Section constructed from photographic mosaics assembled over the course of several years. Because it is a composite of observations gathered as the exposure has developed over time, not all units have been present at all times.

2) Vertical scale as shown. Horizontal scale = vertical in the center of the section only; because of the amphitheater shape of the exposure, there is considerable distortion of scale at edges of the section due to image projection.

the basis of water planes (as constructed from the elevations of their surfaces or their topset/foreset contacts) is another example. In these two cases, a significant change in the elevation of a reconstructed water plane evinces a difference in time. With this in mind it then becomes possible, by means of water-plane correlation procedures analogous to biostratigraphic correlation in bedrock geology, to produce a relative time stratigraphy for a given drainage basin.

STOPS 6, 7 (an optional stop, depending on the quality of the exposure on the day of the trip), **8** and **9** (another optional stop, depending on time) have been included as examples of sets of deposits genetically related to the progressive lowering of the levels of small glacial lakes in north-draining tributary valleys. The various chronological interpretations of these sequences are presented as part of the descriptions for the individual stops. Although the general picture of the depositional setting for these sediments is fairly clear, the details of drainage changes over time, as lower and lower basins became ice-free, are confusing and pose several problems. This is particularly true of the problematic delta system exposed in the Blake pit in Brownfield (**STOP 6**).

The fundamental issue concerning these sequences revolves around the interpretation regarding the activity of ice during their formation. Ideally, the identification of sequences in mapping deglaciation deposits is simply a technique of gathering basic field data. It should not require the prior application of a particular paradigm regarding the physical activity of the associated ice. The deposits at **STOPS 6-9** can therefore be interpreted in two different ways with regard to the activity of ice: one involving downwasting, disintegrating ice; the other involving internally active ice in near proximity. Strictly in terms of the observations from the deposits themselves, neither explanation is in violation of data. However, particularly with respect to **STOP 6**, the first hypothesis involves fewer assumptions, is the simplest, and seems to create fewer problems than it solves. For these reasons, I currently consider it the most valid. I hope that individuals more familiar with the correlation and interpretation of morphosequences than I am will be with us, and that they will consent to lead a discussion.

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FIELD TRIP ROAD LOG

TOTAL TRIP MILEAGE	MILEAGE BETWEEN STOPS	DESCRIPTION
0.00	0.00	Meet at Hiram Falls, overlook, Route 113, Baldwin, Maine.

2.25	2.25	Pass the "Famous Dancemore" hall on the left.
2.45	0.20	Turn right onto Route 117.
3.55	1.10	Cross Saco River.
3.75	0.20	Turn right into Town of Cornish pit.
3.80	0.05	**STOP 1**: Cornish town pit. There are approximately 20' of section exposed here, showing examples of fluvial bedforms. Current directions, as measured at several places in the pit, range from S11W to S85E.
3.85	0.05	Exit pit; turn right onto Route 117. (Note that the road climbs from a lower terrace surface onto the surface exposed at **STOP 1** about 0.1 mi from the pit entrance.)
4.30	0.45	Bear right towards the village of Cornish.
4.60	0.30	Turn right onto Route 25.
4.75	0.15	Turn right next to hardware store.
4.81	0.06	**STOP 2**: Exposure of rhythmically-bedded clay/silt. The composite section exposed here is:

Sand and pebble gravel	(15')
Slightly gravelly medium sand	(7')
Interbedded fine sand, silt, and medium sand; plane-bedded, with soft sediment deformation structures	(9')
Interbedded silty sand and silt with graded bedding	(8')
Rhythmically-Bedded silt and clay/silt; couplets are on the order of 0.1' thick	(11')
Cobble and boulder gravel	(16')

4.87	0.06	Turn left onto Route 25.
6.85	1.98	Pass small pit on right.
7.00	0.15	Cornish-Limington town line.
7.17	0.17	Notice scarp of kame terrace on left.
8.07	0.90	Turn left onto Tucker Road
9.62	1.55	View to left of the Saddleback Hills. Outcrop on right.
9.82	0.20	**STOP 3**: Site of proposed dam for initial stage of Saco valley glacial lake.
11.62	1.80	Bear right through the junction.
12.12	0.50	Four corners. Go straight through.
13.87	1.75	"Ruin Corner". Bear left onto Route 25.
14.86	0.99	North Limington, junction with Route 11. Go straight.
15.64	0.78	Cross Hamlin Brook.
16.64	1.00	Cross Saco River.
16.67	0.03	Turn right onto the River Road.
18.07	1.40	Pass Milt Brown Road on left.
18.57	0.50	Turn left onto Libby Pines Road.
18.64	0.07	Turn right into pit.
18.67	0.03	**STOP 4**: Standish glaciomarine delta. This pit includes an exposure of a topset/foreset contact, whose surveyed elevation is 296' above sea level

(NGVD) (Thompson, Crossen, Borns, and Andersen, 1983). Although the overall quality of the current exposure is not good, the pit has been included as a stop to show the marine sea level indicator nearest the study area. The marine nature of the delta, which at this location is morpholgically a kame, has been established by virtue of an exposure, in the far side of the pit, of Presumpscot Formation clay/silt overlying the granular sediments which constitute the delta foresets.

18.70	0.03	Turn left onto Libby Pines Road.
18.77	0.07	Turn right onto the River Road
22.10	3.33	Turn left onto Route 25; head back toward Cornish.
26.68	4.58	North Limington
28.43	1.75	Pass through "Ruin Corner"
28.78	0.35	Pass Christian Hill Road on left.
30.62	1.84	Cross Back Brook
31.83	1.21	Pass Tucker Road on right.
33.22	1.39	Cornish-Limington town line.
37.87	4.65	Cross Little River in Cornish.
38.27	0.4	Junction of Routes 5 and 25; go straight.
40.52	2.25	Parsonsfield-Cornish town line.
42.36	1.84	Bear left off Route 25 before it crosses the Ossipee River.
45.76	3.4	Take right turn onto Mudgett Road.
46.66	0.9	Right turn onto woods road. Consolidate vehicles here.
47.41	0.75	**STOP 5**: Exposures of "ribbed" moraines.
48.16	0.75	Pick up vehicles. Turn right onto Mudgett Road.
48.56	0.4	Pass pit on left.
49.06	0.5	North Parsonsfield; turn right onto Route 160.
49.46	0.4	Pass elementary school on right.
50.06	0.6	Turn right at Parsonsfield Seminary.
51.96	1.9	Cross Ossipee River.
52.26	0.3	Turn right onto Route 25.
56.86	4.6	Cross over to Route 160, heading north.
57.56	0.7	Cross Ridlon Brook.
57.69	0.13	Turn right at the South Hiram Post Office; continue on 160.
59.09	1.4	Hiram-Porter town line.
61.39	2.3	Enter "The Notch". This large south-southwesterly-trending gorge, which was described by Stone (1899) as having been carved principally by glacial ice, is the local expression of a prominent regional lineament (the bedrock exposed in the gorge is granodiorite, presumably of the Devonian New Hampshire Plutonic Series (Gilman, 1977)). Striations have not yet been found within the notch itself. Although I assume the gorge owes at least some of its current morphology to glacial scour, there is no way of knowing when the scouring occurred. If it took place essentially during the last deglaciation, it could be that the ice retained a considerable degree of internal vigor as it thinned to the point where flow was strongly controlled by very local bed topography. In any event, "The Notch" contained a meltwater system for a

period of time during latest Wisconsin time, and in addition, appears to have served as a spillway for a small proglacial lake.

62.29	0.9	Hiram-Porter town line.
63.69	1.4	Porter-Hiram town line.
66.39	2.7	Pass Burnt Meadow Pond on right.
67.19	0.8	Bear left on Route 160 at the Civil War monument.
67.99	0.8	Bear left at intersection.
69.29	1.3	Turn left at "Patterson's Ponderosa".
70.69	1.4	Turn right just past the Old Blake Farm.
70.89	0.2	Turn left into large, overgrown pit.

STOP 6: Exposure of deltaic sediments.

This pit, which is active on an intermittent basis as a source of road sand for the Town of Brownfield, exposes a section of the foresets of a small delta. The surface slope of the deposit, the direction of dip of the foreset beds, and the position and inclination of meltwater channels on the hillside across the road from the Blake homestead all indicate that the deposit was constructed by a stream, or streams, flowing from the east. The delta is located in the Quint Brook basin, a northward-draining third-order tributary of the Saco.

That the deposit is of glacial origin is assumed, because it is located very close to the drainage basin divide, and I do not believe that the volumes of nonglacial runoff produced from what little basin area remains above the delta could explain the size of the deposit. Presumably, the lake existed because the drainage basin was blocked by ice to the north. For this to be true, an ice margin at the time of sedimentation would by necessity describe a closed loop around the drumlin located 0.5 miles north of the delta, and the lake would be draining over ice, into ice, or along the edge of ice to the north of the drumlin, and out the Shepards River valley towards the Saco. Whether or not such a situation is feasible will surely be a subject of discussion. If the story is essentially correct, it would make a strong case for thinning, stagnating ice in this basin during its deglaciation; if the ice was active and backwasting (which implies that it had a positive surface slope to the north), the lake would be draining <u>into</u> the ice at a point where one would expect the ice to be getting thicker.

71.09	0.20	Back on 'main road'; turn right.
72.19	1.10	Turn right.
72.89	0.70	Cross Cole Brook.
73.19	0.30	Turn right at 'T' junction (New Boston).
74.59	1.40	**STOP 7:** Optional esker exposure.

This is a rapidly evolving exposure, which last year yielded the first unequivocal paleocurrent data from this, the Cole Brook esker system. Those data indicate convincingly that the deposit was constructed from the north, to the south. The bed topography rises to the

south from approximately 580 feet to 830 feet, which is the elevation of the lowest point through which the system that built the esker could exit the Cole Brook basin. There is a lack of collapse features within the core of the esker, which may argue against the idea that the esker was superimposed on the bed from englacial positions. If this interpretation is correct, the the fact that the esker system was constructed in a southerly direction south over north-sloping bed topography indicates that it formed subglacially in a confined tunnel.
The existence of the esker seems to present an argument for relatively thick ice, while the deposits at **STOP** 6 seem to imply thin ice. The simplest explanation for this apparent contradiction would state that the esker formed earlier than the delta (and also, by inference, than the other "non-eskerine" deposits in the Cole Brook and Quint Brook basins). This explanation, involving thinning ice, the development of an "esker phase", followed by the development of a "lacustrine phase", is essentially that proposed by Goldthwait and Mickelson (1982) for the deglaciation of the White Mountains.

74.79	0.20	Turn right
76.09	1.30	View to the left (northwest) of Moat and Presidential Ranges in New Hampshire. Note asymmetrical stoss and lee bedrock hills in the foreground, which is all within Maine.
76.49	0.40	Turn left.
76.79	0.30	Pass "Patterson's Ponderosa" again. Go straight.
77.59	0.80	Junction with West Brownfield Road. Cross road into pit. **STOP 8:** Pit in kame plateau. Exposed are 24' of chiefly medium to coarse sand, with interbedded pebble and cobble gravel near the top of the section. Bedforms include horizontal bedding contained in tabular sets, and climbing ripple-drift. Paleocurrents range from S55E to S90E. The sediments are judged to have been deposited subaqueously by a single meltwater system initiating approximately 2 miles north of the pit. The stream entered a body of open, ponded water choked with relict, disintegrating ice masses. A set of lateral channels, cut in a stepwise manner on the south wall of the Shepards River valley, are believed to have acted as the spillways for this, and several other small glacial ponds. The water levels in these ponds dropped sequentially as lower and lower drainage outlets were exposed upon continuous thinning of the ice in the Shepards River valley.
79.79	2.20	Junction with Route 160, Brownfield. Go straight.
80.89	1.10	Junction with Route 113. Turn left at store and proceed into pit.
80.92	0.03	STOP 9: Exposure in the Brownfield sand plain behind Wally's General Store. Optional stop. General stratigraphy in the section:

<u>Interbedded sand and pebble gravel</u>	(2')
<u>Interbedded sand and silt</u>	(1')
Interbedded medium sand and slightly gravelly sand	(8')

Bedforms exposed at various times have included 0.2' thick tabular sets of cross-stratified sand (lower contacts tangential to set boundaries), and apparently cyclically-bedded sand and pebble gravel. Paleocurrent directions indicate flow from the north and northwest. The deposits are interpreted to have been deposited in a shallow subaqueous environment.

Exposures within the same landform near the village of Brownfield indicate that the deposit was formed from two sources simultaneously: from the north in the Saco valley; and from the west in the Shepards valley. There are two alternate hypotheses which can explain the deposit:

1) All of the sediment was carried by streams fed by glacial meltwaters, implying that there were relict, disintegrating ice masses in the Shepards River valley at a time when the Saco corridor, at least into Fryeburg, was largely ice-free.

2) The western part of the landform was constructed as a nonglacial "fan" very quickly after the Shepards River basin became deglaciated. This explanation can accommodate both the Saco and Shepards River valleys becoming ice-free at quite similar times.

80.95	0.03	Turn onto Route 160; cross Route 113.
81.10	0.15	Cross B&M tracks.
82.40	1.30	Cross Saco River.
83.10	0.70	Pass road to Brownfield Bog on left.
83.90	0.80	Brownfield-Denmark town line.
86.80	2.90	Turn left onto Lake Road.
87.40	0.60	Turn right. Continue on road on a traverse. Note the texture of the sediment (diamicton, exclusively).
88.20	0.80	Turn left.
88.70	0.50	Turn right.
88.80	0.10	Turn right.
88.90	0.10	Turn right back onto Lake Road.
89.10	0.20	Turn into pits.
		STOP 10: Moose Pond kame complex. General features of the stop are described in Section 3.3.2 of the introductory text. This stop is the end of the excursion.

GEOLOGY OF THE EASTERN PORTION OF THE WHITE MOUNTAIN BATHOLITH, NEW HAMPSHIRE

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INTRODUCTION

The Jurassic White Mountain batholith is located in northern Grafton and Carroll Counties, New Hampshire. It is a composite of several overlapping centers of felsic magmatism. Individual centers are strikingly defined by ring dikes and can include plutonic, hypabyssal, and volcanic rock types. Pyroclastic rocks, chiefly rhyolitic tuffs and breccias, are preserved within ring dikes as large subsided blocks. Numerous plutons of subaluminous to peralkaline granites provide an areal continuity to the batholith. The geology of the White Mountain batholith has been mapped by Billings (1928), Billings and Williams (1935), Creasy (1974), Henderson and others (1977), Moke (1946), Osberg and others (1978), Smith and others (1939), and Wilson (1969).

The spatial and temporal geometry of the magmatic centers provides a convenient basis for dividing the batholith into an eastern and a western portion. The eastern portion of the White Mountain batholith (Figure 1) as exposed in the North Conway 15' quadrangle has at least four magmatic centers developed about 175 m.y. ago (Creasy and Eby, 1983). Hitchcock (1878) provided the first outline of the geology of the North Conway quadrangle and certain of his formational names are still in use, e.g. the Conway Granite. However, the work of Billings (1928) remains the chief geological reference for the North Conway quadrangle. Creasy (1974; unpublished maps), Davie (1975), Osberg and others (1978), and Parnell (1975) have provided greater detail to Billings' pioneering work.

GEOLOGY OF THE NORTH CONWAY QUADRANGLE

The geology of the North Conway quadrangle (Figure 1) is summarized in terms of the major magmatic and structural units of the White Mountain batholith.

The Mt. Osceola Granite, a green amphibole ± biotite granite, is the <u>oldest</u> member of the White Mountain magma series exposed in the North Conway quadrangle (Osberg and others, 1978). The number and original extent of plutons of the Mt. Osceola Granite within the North Conway quadrangle is not fully certain due to the complexity and abundance of younger rocks. A whole-rock Rb-Sr isochron for samples from both eastern and western portions of the batholith yields an age of 186 m.y. (Eby and Creasy, 1983) and indicates synchronous intrusion over a broad area . [This age places the Mt. Osceola as the <u>youngest</u> member associated with the large magmatic center that forms the western portion of the batholith.]